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Arctic Ocean Water Mass Balance From Isotope Data

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The distributions of the oxygen 18 and tritium isotopes, and of salinity, yield a residence time of 10 years for the surface and halocline waters of the Arctic Basin. We find a yearly net production of 0.59 m of sea ice and an input of 1.16 m of freshwater from continental river runoff, local precipitation, and Bering Strait salinity deficiency. Using the basin area value with these numbers gives long-term average total net production and transport rates of 0.15 Sv of ice and 0.18 Sv of meteoric component, Bering Strait water not included. If, in addition, a reasonable depth profile of relative current velocity is assumed for the outflow, a yearly input of 2.8 Sv of Atlantic and Bering Strait water is needed to replenish the upper and halocline waters. These numbers should be good to ±20% and are multiyear averages. The isotope data clearly indicate that the Barents Sea is an inflow area for Atlantic water to the basin, but that net export of ice occurs there.

1. INTRODUCTION

The Arctic Basin north of Spitsbergen, including the Barents Sea, is mainly fed by waters from the northernmost Atlantic and, to some extent, from the Pacific through the Bering Strait. Several processes inside the Arctic are of great importance in the water mass balances and time scales and are also of importance for heat budgets and, therefore, the understanding of climate. One is the addition of meteoric water as runoff and as direct precipitation. (Since "freshwater" could be taken to include ice melt, the term "meteoric," or "runoff" is used.) Another is the freezing and melting of sea ice, especially the net balance thereof. Extended efforts have been made trying to untangle and quantify these processes by conventional oceanographic methods including current meters, ice drifters, and using estimates of yearly runoff numbers from rivers in Siberia and the Canadian Arctic. Others have applied chemical information to approach some of these problems [*Tan et al.*, 1983; *Moore et al.*, 1983]. In this paper we are a priori trying to use as little as possible of that previous information and, instead, use the distribution of water isotopes to arrive at independent quantitative information on these processes. We are going to use the oxygen isotopes in the water column in the basin and its outflow to distinguish water sources and tritium to establish time scales.

2. ISOTOPE SCALES AND UNITS

The ratios of the stable isotopes of water, $^2H/^1H \simeq 1.5 \times 1.0^{-4}$ and $^{18}O/^{16}O \simeq 20 \times 10^{-4}$, are each very constant in the bulk ocean. In the hydrological cycle of evaporation and condensation, these ratios change in such a way that they are essentially covariant [*Craig*, 1961; *Dansgaard*, 1964]. Therefore, for this discussion we shall concentrate on one of these isotopes, ^{18}O.

The ratio $^{18}O/^{16}O$ is usually not used directly for comparisons but, instead, the deviation of this ratio from a standard value. For water the standard material is Vienna-Standard mean ocean water (V-SMOW), available for distribution by the International Atomic Energy Agency (IAEA) in Vienna.

The deviation of the $^{18}O/^{16}O$ ratio in a sample, with respect to the ratio in the standard, is expressed as

$$\delta_{\text{V-SMOW}}{}^{18}(\text{sample}) = \left[\frac{(^{18}O/^{16}O)_{\text{sample}}}{(^{18}O/^{16}O)\text{V-SMOW}} - 1\right] \times 1000‰$$

For short this value is denoted δ^{18} in the following discussions.

When ocean surface water evaporates, a considerable isotopic fractionation occurs so that the vapor becomes isotopically lighter (i.e., δ^{18} goes negative). In its global transport toward the north, the atmospheric water vapor undergoes extensive modifications by evaporation, precipitation, and air/sea molecular exchange so that vapor and precipitation in the Arctic has δ^{18} values between −10 and −30‰, varying with time of the year, distance from open ocean, etc. The Atlantic source water, however, shows essentially the unchanged bulk ocean value close to 0.0‰.

Redfield and Friedman [1969] realized that the isotope ratios in Arctic Basin seawater could be used, together with salinity, to separate contributions of meteoric water and ice melt as dilutants and to indicate a net loss of sea ice. They did not have enough geographic and time coverage, however, to make a serious study of Arctic oceanography this way. In 1974, *Vetshteyn et al.* [1974] again made an attempt at using ^{18}O isotopes to identify the freshwater source. *Tan and Strain* [1980] considered that the meteoric δ^{18} value was not well enough known, nor that there were enough data available from the Arctic Basin to warrant a serious study on brine production. We have in the past 5 years collected data on ^{18}O ratios of ice and snow in the Arctic Ocean. Furthermore, a wealth of data on precipitation is indeed available in listings from *International Atomic Energy Agency* (IAEA) [1981]. Assembling this knowledge, we therefore set out to try to use this isotope tool to a fuller extent than had been done before.

As the result of atmospheric testing of fusion bombs, water in nature also contains the radioactive hydrogen isotope tritium, ^3H or T, as HTO. This molecule is subject to isotope fractionation like HDO and $H_2{}^{18}O$, but this effect is negligi-

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ble in comparison with the very large variations by time and space caused by the unique spike-type tritium source function. See, for instance, *Östlund* [1982] for a more comprehensive description. Tritium is expressed in TU, where 1 TU stands for a T/H ($^3H/^1H$) abundance ratio of 10^{-18}.

3. Mass Balance Equations

Consider a sample of seawater collected at some depth in the Arctic Ocean or its outflow, having a salinity value of S and a δ^{18} value of X. This water can be considered having been formed in the following way:

F_a kg of original Atlantic water with salinity S_a, and δ^{18} value X_a, is mixed with F_r kg of continental runoff and local precipitation with salinity S_r (=0) and δ^{18} value X_r. Now also add to this mixture F_i kg of meltwater from ice with salinity S_i and δ^{18} value X_i. The mass balance equations will then look as follows:

$$F_a + F_r + F_i = 1 \tag{1}$$

$$F_aS_a + F_rS_r + F_iS_i = S \tag{2}$$

$$F_aX_a + F_rX_r + F_iX_i = X \tag{3}$$

It shall be noted that negative F_i values will be obtained if ice has formed and been removed from the mixture. These equations will thus model dilution by runoff and the net effect of ice melted or ice formed on a shelf. This process for "brine" production has been observed in nature by *Melling and Lewis* [1982]. Included in F_a is subsequent entrainment of more Atlantic water during the transport of the mixture to the position in the water column where we collected the sample. The total amount of each component in a water column will be obtained by integrating the fractions by depth.

In an attempt to establish time scales for the water masses in the Arctic Ocean, *Östlund* [1982] used a tritium/salinity age concept. This approach required linearity between salinity and tritium concentration through a substantial depth interval and assumed that the salinity deficiency was entirely due to meteoric water (i.e., it did not take into account the formation or melting of ice). The tritium value of the freshwater diluting the Atlantic waters was obtained by extrapolation of the tritium/salinity line to zero salinity. This value was then matched with an empirical source function of the tritium history in Arctic precipitation and runoff 1957–1980. We are now modifying this concept so that when F_r is found, the tritium concentration, T_r, for runoff is calculated from the tritium mass balance equation (4) below, where T_a and T_i are tritium values of Atlantic source water and ice, respectively, and T is the tritium value measured in the mixture:

$$F_aT_a + F_rT_r + F_iT_i = T \tag{4}$$

Again, the T_r value will be matched to the tritium runoff source function to obtain runoff year, "vintage." This way, addition of meltwater or salinity increase by freezing is properly accounted for, but the transfer times obtained this way generally differ by no more than ± 1 year from those obtained in the 1982 paper. This is due to the very steep ($\sim 30\%$ per year) slope of the tritium source function at the critical time period 1965–1971. A weakness of this "freight car age" will be discussed below.

Another way to determine the residence times of Arctic waters was reported by *Östlund et al.* [1982]. This method requires the assay of tritium and its radiogenic daughter ^3He of the sample, and it yields a seal-off time for the final mixture, regardless of composition, and without the use of the tritium runoff source function. Special procedures needed for ^3He sampling have generally not been available to us so that data base is limited.

Bering Strait water has δ^{18} and salinity values (~ 32.4) that make the water look somewhat like Atlantic water diluted with meteoric water. Our model will thus, at this stage, consider freshwater equivalent of the salinity deficiency of that water to be included in F_r.

4. Salinity and Isotope Data

In equations (1)–(3) above, the source values have to be known, and we have used the following rules:

1. Salinity: The Atlantic source water has a salinity of $S_a = 34.92$, meteoric water $S_r = 0$, and ice $S_i = 4.00$.

2. ^{18}O: The δ^{18} value of the Atlantic water is rather well established from previous measurements, and it can also be obtained by averaging a large number of our measurements on Arctic waters at "full Atlantic salinity," usually below about 300 m, and by extrapolating the δ^{18}/salinity relationship to pure Atlantic water, $S = 34.92$. The best value is $X_a = 0.3‰$ with an error of no more than $\pm 0.1‰$.

To find the runoff source value X_r, one shall not use the technique of linear extrapolation of δ^{18}/salinity to $S = 0$, as it would discount the existence of two freshwater sources. Instead we made a literature search of all available isotope data for precipitation north of 60°N. Such data are available in publications by *IAEA* [1981]. In these listings, western and coastal Arctic data are over represented. Since tritium information is geographically more evenly spread, we applied a standardization technique similar to the one used by *Östlund* [1982] for tritium. The weighted δ^{18} value obtained for inland precipitation in the relevant areas is $X_r = -21 \pm 0.7‰$ (1 standard error). *Krouse and Mackay* [1971] found $-20.2‰$ in the lowest part of the Mackenzie River. Other data are available in a paper by *Bédard et al.* [1981], pointing at $-21.1 \pm 0.6‰$. From data on deuterium ratios in Eurasian rivers by *Soyfer et al.* [1967], one would get δ^{18} values from -15 west of Ural to $-23‰$ farthest east. With this information in hand, we decided to use an X_r value of $-21‰$, and we also made sensitivity tests at -19 and $-23‰$ on almost all cases. Individual snowfalls and rains vary much more. We found values in fresh snow from $-29.9‰$ at Fram 3 (ice camp in late winter) to $-11‰$ on a ship in open water along the ice edge in late summer. The very low values -30 to $-55‰$ seen in Antarctica and on top of the Greenland ice cap are not representative of low altitude precipitation relevant to our investigation. When ice is formed by freezing under equilibrium conditions, there is a slight isotope fractionation in that the solid phase will have an isotope value 3‰ higher than the liquid phase [*O'Neil*, 1968]. The origin and isotopic value of any ice, now residing at a sampling station, could be quite different from that of the water column due to the difference in transport patterns of ice and liquid water. Considering that sea ice is not really freezing under true equilibrium conditions, we have used the following rules for selecting each X_i value: It is the δ^{18} value of the surface or shallowest available sample in the pertinent column of water, plus 1.5‰. Even if this may be rather close to the truth during freezing, during melting of ice, there would certainly be no fractionation whatsoever. In this paper we will not address this asymmetry.

For the tritium ratios T_a and T_i, we have applied the equivalent criteria for fractionation. However, the time information deduced from the tritium data is not very sensitive to small variations caused by this effect.

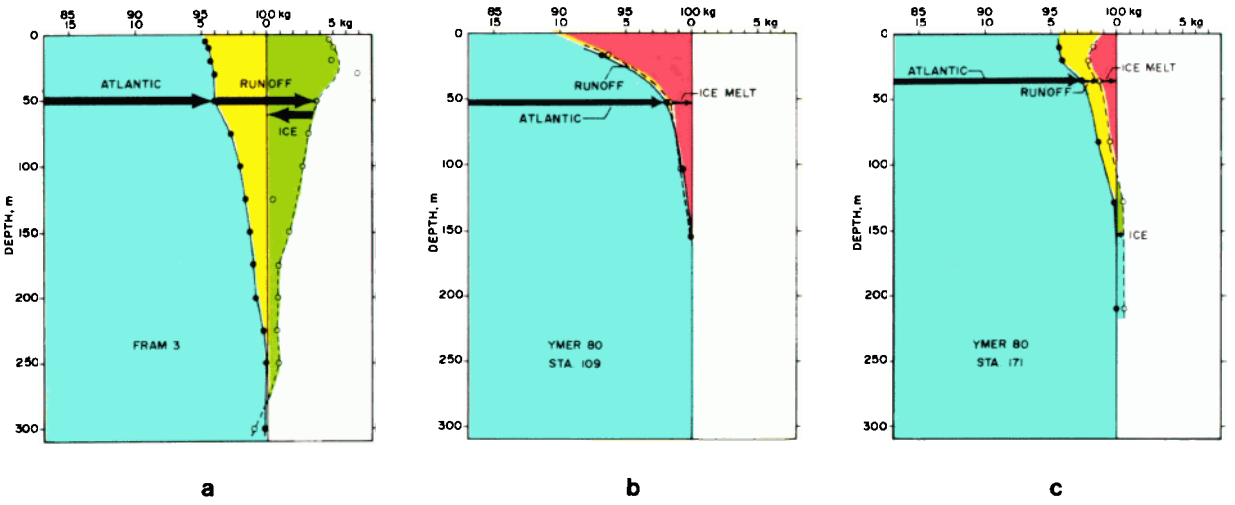

Plate 1. The makeup of water at depth 0–300 m at three stations: (*a*) East Greenland Current, outflow; (*b*) West Spitsbergen Current, inflow; (*c*) mid Fram Strait.

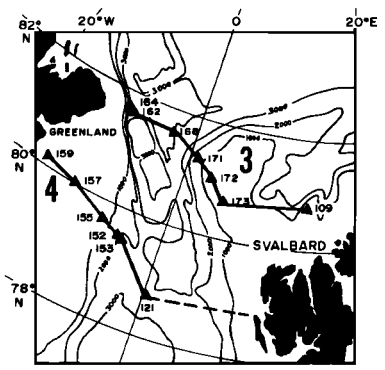

Plate 2. Location of Fram Strait sections.

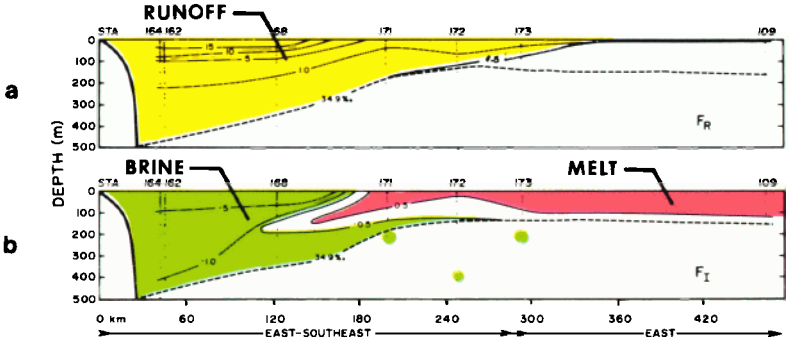

Plate 3. YMER 80 section across Fram Strait at about 81°N.

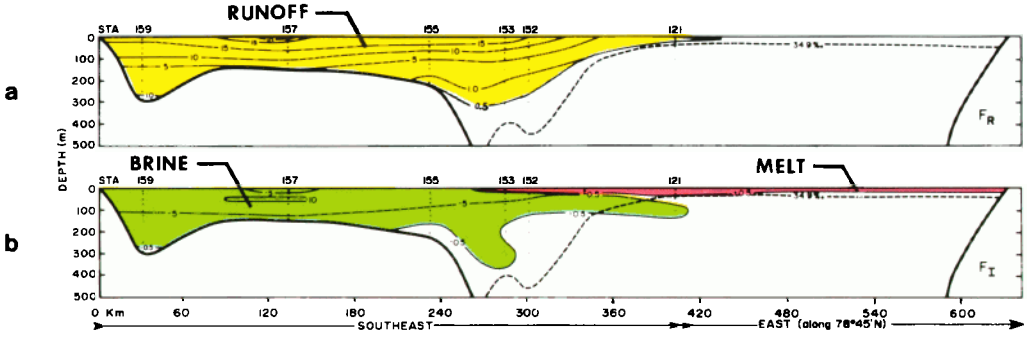

Plate 4. YMER 80 section across Fram Strait at about 79°N.

TABLE 1. Tritium in Arctic Runoff

Year	TU	TU81N
1953	21	5
1954	225	54
1955	131	33
1956	196	53
1957	175	49
1958	475	141
1959	573	180
1960	380	127
1961	430	152
1962	1120	418
1963	2739	1083
1964	2421	1013
1965	1645	728
1966	1111	521
1967	710	351
1968	492	258
1969	397	221
1970	326	192
1971	341	211
1972	258	169
1973	222	155
1974	184	135
1975	151	118
1976	119	98
1977	95	82

TU values are T/H ratios ($\times 10^{18}$) in June as of each year; TU81N are values age-corrected to 1981/01/01, see text.

5. Sampling and Measurement Methods

On ships or ice stations, water samples were usually taken by standard Niskin bottles, and 1 liter of water transferred to bottles previously cleaned and filled with argon gas. This technique, described by *Östlund et al.* [1974], serves to protect the samples from contamination by atmospheric humidity. In most cases, salinities were determined by the chief investigators of the cruise or ice station. Considering the many different cooperating institutions and scientists, we claim that the salinity values are not better than ± 0.010 to ± 0.015. That limits the interpretation only at salinities very close to Atlantic values, where the errors in isotope ratios are comparable to natural scatter anyway.

The analysis of $^{18}O/^{16}O$ in water is performed by equilibrating 1 ml of water sample with CO_2 of known isotopic composition by shaking for 2 hours at 33°C, a method originally proposed by *Roether* [1970]. This equilibration is carried out on batches of eight unknown water samples and 2 aliquots of standard water. The gas phase CO_2 is then analyzed mass spectrometrically.

According to an international agreement in 1976, we refer our data to the primary standard V-SMOW. Some researchers may still refer to the older SMOW-standard. The difference between a value in that scale and the V-SMOW scale is very small [*Gonfiantini*, 1978]:

$$\delta^{18}_{\text{V-SMOW}}\ (\text{SMOW}) \simeq +\ 0.07‰$$

Mass spectrometric analyses were performed on a Micromass 903 spectrometer. The error in the reported δ^{18} values is less than 0.1‰.

Tritium was determined by electrolytic enrichment and low-level gas proportional counting as described by *Östlund et al.* [1974]. The measurement errors are ± 0.06 TU or $\pm 3.5\%$, whichever is greater. In order to compare tritium values for samples collected at various times, all tritium ratios were recalculated to the TU value at the reference time 1981/01/01

Fig. 1. Geographical distribution of samples. Open circles mark stations with vertical profiles. L is for LOREX 1979, F1 and F3 are FRAM stations 1979 and 1981, connected circles YMER 1980, numbered circles submarine samples 1978, filled triangles single samples by submarine 1981, and filled circles incomplete profiles from NPI cruise 1980.

TABLE 2. Data Obtained on Fram 3

Sample	Depth, m	Temp, °C	Salinity, ‰	Sigma Theta, σ_θ	Tritium, TU81N	δ^{18}, ‰	F_a, kg	F_r, kg	F_i, kg	T_r, TU81N	T/S, years	T/He, years
312	5	−1.82	33.083	26.623	20.51	−1.65	95.25	9.44	−4.69	182	10	4.5
112	10	−1.83	33.173	26.696	19.99	−1.64	95.54	9.41	−4.95	177	9	4.5
311	20	−1.83	33.212	26.728	20.09	−1.60	95.64	9.22	−4.85	182	10	4.7
310	30	−1.83	33.247	26.756	19.05	−1.92	95.96	10.84	−6.80	148	8	
612	50	−1.78	33.391	26.872	20.83	−1.28	96.01	7.64	−3.65	226	12	
110	75	−1.78	33.826	27.226	16.74	−0.91	97.19	5.87	−3.06	221	12	7.8
308	100	−1.70	34.069	27.421	15.38	−0.70	97.84	4.86	−2.70	237	12	7.5
307	125	−1.76	34.304	27.614	10.78	−0.15	98.25	2.14	−0.39	301	13	
306	150	−1.52	34.369	27.660	9.14	−0.33	98.58	3.06	−1.64	164	9	
305	175	−0.64	34.511	27.742	9.65	−0.09	98.89	1.88	−0.77	285	13	12.7
108	200	−0.20	34.587	27.783	8.03	−0.05	99.11	1.69	−0.80	220	12	12.5
304	225	0.98	34.780	27.870	6.63	0.09	99.65	1.03	−0.67	223	12	8.6
303	250	1.13	34.851	27.917	5.88	0.11	99.87	0.94	−0.81	165	8	8.6
301	300	1.06	34.905	27.965	4.97	0.47	99.81	−0.86	1.06			11.7
607	350	1.72	34.943	27.948	4.98	0.17	100.13	0.66	−0.79			
606	400	1.51	34.974	27.989	4.97	0.27	100.16	0.16	−0.32			
605	450	1.23	34.945	27.986	4.64	0.37	100.00	−0.35	0.35			
107	500	0.88	34.968	28.028	3.47	0.31	100.12	−0.04	−0.07			
603	750	0.48	34.968	28.053	2.50	0.22	100.18	0.41	−0.59			
106	1000	−0.17	34.976	28.096	1.74	0.00	100.34	1.52	−1.86			
512	1250	−0.37	34.962	28.095	1.55							
105	1500	−0.60	34.968	28.110	0.82	0.32	100.11	−0.09	−0.02			
510	1750	−0.68	34.953	28.102	0.62	0.19	100.15	0.56	−0.71			
507	2500	−0.82	34.930	28.089	0.27							
102	3000	−0.80	34.906	28.069	0.78	0.20	99.99	0.50	−0.49			
101	3502	−0.76	34.939	28.094	0.12	0.23	100.07	0.35	−0.43			
501	4186	−0.69	34.980	28.124	0.05	0.33	100.14	−0.14	−0.00			

F_r and F_i values are highly unreliable at salinity > 34.85 (i.e., below 300 m).

using the new half-life 12.44 years and referring directly to the U.S. National Bureau of Standards Standard Reference Material 4926. We call this tritium value TU81N, for new half-life. In the paper by *Östlund* [1982], an age adjustment was made to 1980/12/31, using the old half-life 12.26 years. The "TU80" values and the TU81N values, even if referring to almost the same time, differ so that TU81N = 1.0307 × TU80. For a discussion on tritium scales and half-life, etc., refer to *Mann et al.* [1982].

In Table 1 are listed the age-corrected tritium values, TU81N, of Siberian and Canadian runoff to the Arctic Ocean, converted from the TU data in the 1982 paper, first column. The T_r value calculated from (4) is matched to this table to find the "vintage year," and the difference between that year and year of sampling expresses the length of time as the "freight car" age, the time apparently elapsed between admixture of runoff on the shelf and the time of sampling, somewhere else in the basin.

6. Experimental Result

For our purpose it is desirable to have a large number of samples spread over the Arctic Basin and also to have good coverage of the outflow. Thanks to good cooperation from colleagues and agencies, we now have ^{18}O and tritium data from stations as indicated in Figure 1.

Space does not allow us to present all data in tabular form, but we shall present raw data at one station at the beginning of the outflow from the basin, with good depth coverage, namely, Fram 3, Table 2. For choice of component parameters, see section 4 above. This station exhibits large negative F_i values (i.e., loss of water due to freezing). An example may best explain the result: Sample 612 collected at 50 m depth, with a salinity of 33.391, has a measured δ^{18} value of −1.28‰. If the salinity deficiency (4.4% freshwater by weight) had been caused by runoff water only (at $\delta^{18} = -21$‰), the δ^{18} value of the mixture would have been −0.64‰, not −1.28‰. There is thus a very clear indication that ice has been involved.

The mass balance equations (equations (1)–(3)) yield the fraction of each of the components, and this is to be understood as follows for this sample; refer to Table 2 and Plate 1*a*: 100 kg of water sample 612 has been formed by first mixing 96.01 kg of undiluted Atlantic water (blue, in the figure) with 7.64 kg of runoff to make 103.65 kg, illustrated as a yellow field from 96.01, past the 100 kg line to the 103.65 point. From that mixture 3.65 kg of ice, a mixture of Atlantic (blue) and runoff (yellow) has been removed by freezing, thus forming a green field between the 100 kg line and the 103.65 kg point. This process has presumably occurred on a shelf somewhere, and that ice is no longer at the same geographical location as the water sample; it may even have left the basin already, and the remaining quantity of seawater is 100 kg. When matched to Table 1, the TU81N value 226 places the runoff vintage anywhere from 1968 to 1971, so the time elapsed between runoff since 1969.5 ± 1.5 and date of sampling, 1981.5, is 12 years. Since this water is just to leave the basin, the residence time since initial formation on the shelf is 12 ± 1.5 years. (Unfortunately, some Chinese bomb tests caused a wiggle in the tritium source function just around 1971.)

By increasing depth, when the salinity approaches 34.9 and the δ^{18} value comes close to that of the Atlantic, the relative uncertainties of the calculated F values increase rapidly, and our model breaks down. Also, no vintage can be estimated from the tritium number. However, there is still a structure of tritium by depth reflecting the tritium profile of the original Atlantic water as it entered the Arctic Basin.

A station in the inflow to the Arctic Basin will present a different picture, and we select a station from the Swedish icebreaker expedition, YMER 80, in August of 1980 [*Anderson*

TABLE 3. Runoff and Ice Melt or Ice Production

Station	X_i, ‰	R, m	I, m	Age, years	Integration depth, m: R	Integration depth, m: I
			Fall Patrol, 1978			
1	−0.50	7.8	−3.8	10	122	122
2	1.20	1.0	+1.1	12	122	122
3	1.50	1.3	+1.5	13	122	122
4	0.00	8.8	−3.5	11	166	166
5	−1.00	13.3	−3.8	10	166	166
6	−1.40	10.4	−2.7	10 ± 1	166	166
7	−2.5	18.9	−6.7	10	166	166
8	−1.7	19.0	−5.4		166	166
9	−1.3	17.0	−5.5		166	166
			Fram 1, 1979			
18	−1.0	18.9	−8.9	11	292	292
30	−1.0	18.2	−8.8	11	291	196
65	−1.0	14.5	−7.6	11	297	161
			LOREX, 1979			
1 + 2	−1.0	10.4	−3.7	8 ± 1	226	155
			Barents Sea, 1980 Scattered Locations			
		1.0	+1.0	NC	···	···
			YMER-80, 1980			
102	1.0	1.0	+1.0			
104	1.0	1.0	+1.0	NC	200	200
105		2.1	+1.0			
109	1.0	0.5	+3.2	NC	150	150
110		1				130
111		1				30
112		1				22
117		1				77
119		1				50
121		0.3	+1.4	NC		50
142		1	+0.3	NC		50
151		1	+0.5	NC		50
152	1.0	11.7	−4.9	11 ± 1	205	205
153	1.0	12.5	−5.9	9 ± 1	207	207
154		10	NA		204	204
155	1.0	13.5	−7.8		214	214B
157	1.0	15.5	−9.8		145	145B
159	1.0	17.3	−9.1		290	290B
162	−2.0	15.5	−6.9	12 ± 1	249	197
164	−2.0	16.9	−9.0		350	200
168	−2.0	13.3	−6.5	9 ± 1	306	306
171	1.0	2.0	+1.1	8 ± 1	209	81
172	1.0	1.5	+0.3	8 ± 1	156	156
173	1.0	1.6	+1.3	NC	···	···
176	1.0		0	NC	···	···
185	1.0	0.3	+0.4	NC	126	126
191	1.0	0.5	+2.0	NC	208	208
207		1				153
209	1.0	0.4	+1.6	NC		76
211						
215	1.0	1.4	+1.5		154	51
			Fram 3, 1981			
1	−1.0	10.7	−5.5	12 ± 1	250	250

Cf. map, Figure 1, for locations. R is equivalent depth of column of runoff water, I is that of ice melt, if positive, and of water removed as ice, if negative. B = bottom depth.

and Dyrssen, 1981], which gave us a large number of samples. In Plate 1*b* we plotted data for YMER 80 station 109, which is just north of Spitsbergen in the extension of the West Spitsbergen Current. The Atlantic water from the south can have only local precipitation (yellow) and ice melt (red) as dilutant. Here, the meteoric input is very small, ~0.5%, and the only significant freshwater source is ice melt.

YMER station 171, cf. Plate 1*c*, is located near the front between East Greenland Current outflow and West Spitsbergen Current inflow. In the upper layer, any early ice formation has become more than compensated by addition of ice melt; below 100 m the data indicate, however weakly, some brine addition.

The YMER expedition yielded two good sections across the Fram Strait (see map, Plate 2). In the two sections across the strait, Plates 3 and 4, the isopleths mark the amounts of the components to make 100 kg of resultant water. As in previous figures, presence of runoff is yellow, of ice melt is red, and of "brine" (i.e., loss as ice) is green. The uncertainty of our measurements, and natural source fluctuations, prompt us to omit color in fields with less than 0.5% of the component. The figures clearly show that ice melt is the dilutant in the eastern part, the inflow, of the Fram Strait. The presence of meteoric water (i.e., river water) also carries with it "brine" (i.e., loss of ice due to freezing on the shelves of the Arctic Basin) in the East Greenland Current outflow.

7. Comments to Obtained Data

7.1. *Runoff and Ice*

In Table 3 we have summarized data from all stations where pertinent data are available. To obtain the height of column of runoff, R, and of ice melt, I, the F_r and F_i values were smoothed with a spline function and F values integrated by depth down to a level where salinity and δ^{18} values are indistinguishable from Atlantic values. This usually occurs at, or slightly above, temperature maximum, if there is any, or at 200–300 m depth. The Fall Patrol submarine sampling did not go quite deep enough for that criteria. To account for that, the F_i and F_r values were extrapolated below deepest sample depth, using the trend of the F values and the pattern at similar stations (Fram and LOREX). The error caused by this extrapolation is less than other uncertainties in the final result. Stations 8 and 9, in the Canada Basin, present special problems since the tritium/salinity line points at a seawater source value of $S_a = 33$ instead of 34.9, indicating Bering Strait water as source [cf. *Östlund*, 1982, Figure 7]. We still use 34.9, thereby including the salinity deficiency as meteoric water.

In the Barents Sea and at the YMER stations in the West Spitsbergen Current extension, both the tritium and the ^{18}O data show essentially zero runoff addition, and any salinity deficiency below a few meters depth is clearly ice melt. This indicates absence of major rivers discharging upstream of these areas. It also tells us that Barents Sea is mainly transporting water from the Atlantic to the Arctic Basin, even if ice transport is directed out from the Basin.

7.2. *Tritium "Ages"*

The tritium/salinity age described above is implicitly a "freight car age," really valid only if no mixing took place in the Basin. However, it is quite obvious that the waters are indeed quite well mixed horizontally. We have tried to use a straightforward continuous mixing model to describe the tritium distribution, but that has failed; it does not produce the high tritium number found around 1978 [*Östlund*, 1982]. The tritium/helium age is a somewhat better approach, generally giving only slightly lower ages. A more sophisticated modeling effort will have to be tried, but we doubt that the age scale will change by that effort. In the cases where F_r becomes small ($< \sim 0.5\%$), no meaningful age can be derived. In Table 3 the mean age and the standard deviation (not standard error) of the distribution are listed for stations with many samples of calculable age. NC means that no age could be calculated.

TABLE 4. Water Mass Balance Based on Basin Area Averages

	Runoff R	Ice I	Units
Column height of components	11.6 ± 0.7	5.9 ± 1.2	tons/m^2 (m)
Yearly average production	1.16 ± 0.07	0.59 ± 0.12*	m/year
Basin area (excluding Barents Sea)	8.0		M km^2
Total outflow	0.29 ± 0.03	0.15 ± 0.03	Sv
Bering Strait "freshwater"	0.11 ± 0.03	···	Sv
Net meteoric water flux	0.18 ± 0.04		Sv

*Liquid water equivalent.

8. Oceanographic Implications

8.1. *Time Scales*

Stations Fram 3 and YMER stations in the East Greenland Current can be considered representative for the surface and halocline waters leaving the Arctic Basin. Also Fram 1, even if farther north, could be included. The tritium age for the waters leaving here was 11 years in 1979 (Fram 1), 10 ± 2 in 1980 (YMER), and 12 ± 1 in 1981 (Fram 3). The 11 ± 1 year average of these represents the average time the runoff component spends on its journey from river mouths to basin outlet. The residence time for the Atlantic source water in the same strata should be the same, within a year, since mixing and freezing should happen each year. We thus claim that the long-term average residence time, based on tritium/salinity, for waters leaving the basin in the East Greenland Current is 11 ± 1 years. As will be shown below, this does not hold for ice, which seems to have a much shorter dwelling time.

We will now consider the "tritium age" of samples inside the basin, Fram 1, LOREX, Fall Patrol, T3. That average is 10.0 years with a standard deviation of 1.5 years. The upper waters in the Arctic Ocean thus show almost the same age as the outflow, and the horizontal mixing must be occurring on a time scale much shorter than the residence time. The same residence time of about 11 years was obtained from the first sampling in 1972 (GEOSECS in *Östlund* [1982]), when T_r must have been above 800 TU, through 1983 with T_r value around 200, both in the age-corrected scale. The sharp peak of the tritium source function in 1962–1964 will cause high estimates of the residence time in the first 10–15 years after that. We thus somewhat arbitrarily take 10 ± 1 years (maximum error) as the long-term average residence time for the Arctic Basin surface and halocline waters. A residence time of 10 years was estimated from river discharge data by *Aagaard and Coachman* [1975] (i.e., based on a completely different type of information).

8.2. *Water Masses Balance*

To estimate reliably the total amounts of the various water masses that form the upper layers of the Arctic Basin, we would like to have a large number of vertical profiles evenly distributed over the basin area. We are not that fortunate, but will try to make do with what we have. To avoid putting too much emphasis on the Fram Strait area, we form averages using data from Fall Patrol, Fram 1, LOREX, YMER 162, Fram 3, and Spring Patrol, cf. Figure 1 for locations.

The average of "depth of water column" of runoff and of ice formation (i.e., the R and I values (negative) in Table 3) were calculated and listed in the first row of Table 4. These values, $R = 11.6 \pm 0.7$ m and $I = -5.9 \pm 1.2$ m, are the quantities of water per unit basin area, that have been involved in forming the presently found isotope/salinity pattern represented by the samples. With our 10 year residence time, the average net yearly production is thus one tenth that. Note that the number reflects the net balance of ice production in the Basin as a whole. We have no means to find how large the debit and credit posts are (i.e., how much freezes each winter and how much melts each summer). As we have no means to distinguish Bering Strait "freshwater" contribution from runoff, that former component is included in the runoff quantity.

According to these results, the "image" of 5.9 m of water lost as ice is present in the waters of the basin. If the average thickness of ice over the basin is, say, 3.0 m, the residence time of the ice would be about 5 years. Clearly, our estimate of ice residence time is directly dependent on the assumed average thickness: our production value 0.59 m/year (water) is not, because it depends exclusively on isotope and salinity data.

To convert from depth of water column per year to water mass transport, we only need the area of the basin. Our isotope data on samples from a cruise by the Norsk Polarinstitutt (NPI) and the YMER expedition indicate that the Barents Sea and the area just north of there contain only Atlantic inflow, so these areas shall not be included in our basin area. Furthermore, on the shelf areas, the water spends a relatively short time, 1–2 years, so for effective area we will use the value 8.0×10^6 km^2 and obtain the total outflows of 0.29 Sv of runoff plus Bering Strait "freshwater," and 0.15 Sv of ice, cf. Table 4. These numbers should be good to within $\pm 20\%$. *Wadhams* [1983] has recently estimated 0.10–0.15 Sv of ice transport through Fram Strait based on thickness and velocity of the ice. His value is based on a short-term average which should not necessarily be the same as our long-term (10 year) average, which includes the entire ice export, not only through Fram Strait.

Our model does not address the question of where the outflow is; our numbers are total fluxes and should thus include the sum of the East Greenland Current and the outflow through the Canadian Archipelago.

For the Bering Strait inflow there are good transport estimates available [see *Aagaard and Greisman*, 1975], 1.5 Sv (± 0.5) at salinity 32.4 which would corrrespond to a "freshwater" input of 0.11 ± 0.03 Sv. The net, true meteoric water input, runoff plus local precipitation, should thus be 0.18 Sv. This number is higher than the 0.1 Sv of *Aagaard and Coachman* [1975] based on river discharge estimates.

8.3. *Outflow Estimates*

We have two more or less complete sections across Fram Strait on the YMER expedition at about 79°N and 81°N, cf.

TABLE 5. Water Mass Balance Based on Outflow

Total Transport	Sverdrups 1.5	2.9	3.6	Basin Calculation
Runoff (including Bering Strait freshwater) R	0.15	0.29	0.36	0.29
Water deficiency (ice export) total I	0.065	0.14	0.16	0.18
Ice export via Barents Sea		0.04*		

Data from YMER stations 162 and 168 in Fram Strait transports of runoff and ice for several values of total water outflow, East Greenland Current plus Canadian Archipelago.

*To make up to match 0.18 obtained in the basin calculation.

Plates 3 and 4, from which we can also try to find the total transport rate. Here we have to make more assumptions than in the previous discussion, and the results become more uncertain. We must first assume that the sections are representative of the long time average situation in the Fram Strait, which may or may not be true. Since there is a pronounced depth gradient in the F_i and F_r values, and presumably also in the current velocity, we should not multiply average F values with total water transport to obtain flux. We assume a depth of halocline outflow of 200 m and divide the depth into four 50-m slices, each quarter transporting a fraction of a total water quantity of (W_j) so that

$$\sum_{j=1}^{4} (W_j) = 1 \tag{5}$$

For each depth range, the fractions of runoff $(F_r)_j$ and ice equivalent $(F_i)_j$ and the fluxes of runoff (Q_r) and the ice equivalent (Q_i) are

$$Q_r = Q \times \sum_{j=1}^{4} (W_j) \times (F_r)_j \tag{6}$$

$$Q_i = Q \times \sum_{j=1}^{4} (W_j) = (F_i)_j \tag{7}$$

with Q the total outflow, surface to 200 m.

For (W_j), the relative fluxes, we select the values 0.40, 0.25, 0.20, and 0.15, respectively, as used by *Anderson et al.* [1983], and apply this scheme to YMER stations 162 and 168 (cf. Plate 3). We do not use width of current, or transport, but search for the Q value that best matches the Q_r and Q_i values, 0.29 Sv and 0.18 Sv, found in our basin production calculations above. The values for these different choices of Q are listed in Table 5. We would prefer $Q = 2.9$ Sv for total liquid water outflow through the East Greenland Current plus the Canadian Archipelago, since it matches the fluxes of runoff found in the basin calculation. Also the deficiency in ice transport, 0.14 Sv instead of 0.18 in basin computation, would support the concept of an export of ice via Barents Sea of 0.04 Sv, which does not contradict the ice melt numbers we get in that area.

We thus find the best transport rate for total outflow from the Arctic Basin to be 2.9 ± 0.4 Sv of liquid water, including surface and halocline waters, and 0.18 Sv of sea ice. This would require an inflow of 2.8 Sv of Atlantic (and Bering Strait) water. On transports of Atlantic water to form deeper and bottom waters this study yields no information.

This outflow calculation is considerably more speculative and uncertain than the basin calculation. It is also realized that, owing to entrainment from the inflow, transport data from the East Greenland Current are difficult to assess. We note that *Stigebrandt* [1981], in his model study, uses 2.75 Sv as one of his preferred transport numbers, and he also arrives at ice export values of 0.08–0.12 Sv through the East Greenland Current plus Canadian Archipelago. The result on ice transport by *Wadhams* [1983] quite nicely brackets our value of 0.14 Sv for the Fram Strait ice export.

9. Conclusions

Our estimates of the water mass balance, and residence time of the upper waters, are based only on isotope and salinity data and area of the basin. We find a residence time of 10 ± 1 years for the halocline waters. Our results show the net yearly average ice production to be 0.59 m per year and the meteoric component (rivers, direct precipitation, and Bering Strait) to be 1.16 m per year. Using a basin area of 8 M km^2 (excluding the shelves) and accounting for the Bering Strait inflow, these numbers are equivalent to 0.18 Sv of true freshwater and 0.15 Sv of ice. If we also assume only a reasonable relative depth distribution of current velocity in the Fram Strait and let it represent all outflow, we get a yearly average transport of 2.8 Sv of Atlantic water in the halocline waters. The results truly represent the long-term averages because they are based on tracer distributions.

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